

Mingling and mixing features in basaltic andesites of the Eastern Cordillera (Central Andes, 24°S): a petrographic and microanalytical study

ANNA GIONCADA^{1*}, NATALIA HAUSER², MASSIMO MATTEINI³, ROBERTO MAZZUOLI¹, RICARDO OMARINI²

¹ Dipartimento di Scienze della Terra, Università degli Studi di Pisa, Italy

² Facultad de Ciencias Naturales, Universidad Nacional de Salta, Argentina

³ Università di Brasilia, Brazil

ABSTRACT. — The basaltic andesites of the Diego de Almagro volcanic complex (Central Andes, 24°S, 66°W) were erupted during Late Miocene at 300 km from the active arc, in the Eastern Cordillera, at the easternmost end of an important strike-slip structure, the Calama-Olocapato-El Toro. These volcanic rocks preserved traces of their pristine history in the form of xenocrysts and in the diverse populations of phenocrysts coexisting due to magma mixing. In Central Andes, crustal contamination commonly conditions the composition of magmas ascending through the thick continental crust, masking the characteristics of the mantle sources. Thus, we carried out detailed petrography and mineral chemistry studies on selected Diego de Almagro basaltic andesites, with the aim to identify the populations of crystals in equilibrium with mafic melts. The results suggested that mixing of basaltic and andesitic magmas occurred during the ascent and storage in the crust. The fraction of basaltic magma prevailed in the mixing process compared to the andesitic one. Crystal zoning and disequilibrium features indicate that magma mixing produced intimate hybridisation. The preservation of primitive crystal assemblages is probably due to the geological situation in which the Diego de Almagro complex emplaced, that favoured the formation of monogenetic volcanic centres rather than of a central volcanic edifice, thus limiting the residence of mafic magmas in the crust and their interaction with crustal partial melts.

RIASSUNTO. — Le andesiti basaltiche appartenenti al complesso vulcanico Diego de Almagro (Ande Centrali, 24°S, 66°W) sono state emesse nel Miocene Superiore a ben 300 km verso E dall'arco attivo, nella Cordigliera Orientale, all'estremo orientale di una delle più importanti strutture trascorrenti trasversali andine, la Calama-Olocapato-El Toro. Le lave e le piroclastiti di questo complesso vulcanico conservano tracce della loro storia evolutiva sotto forma di xenocristalli e di popolazioni di cristalli in disequilibrio che coesistono grazie a processi di mixing di magmi a diversa composizione. Nelle Ande Centrali, la contaminazione crostale condiziona fortemente la composizione dei magmi che risalgono attraverso la spessa crosta continentale, mascherando le caratteristiche della sorgente mantellica. Per questo motivo, abbiamo ritenuto interessante effettuare uno studio di dettaglio della petrografia e della chimica dei minerali delle andesiti basaltiche di Diego de Almagro, allo scopo di identificare le popolazioni di cristalli in equilibrio con i magmi più primitivi. I risultati raccolti finora hanno mostrato che le andesiti basaltiche sono il prodotto di un processo di mescolamento di magma basaltico ed andesitico verificatosi durante la risalita del magma nella crosta, in condizioni pre-eruttive. Nel processo di mixing che ha prodotto le andesiti basaltiche la frazione di magma basaltico era prevalente in volume rispetto a quella di magma andesitico, e le proprietà fisico-chimiche dei due magmi permisero la realizzazione di una avanzata ibridizzazione. L'associazione mineralogica primitiva si è potuta preservare grazie

* Corresponding author, E-mail: gioncada@dst.unipi.it

alla particolare situazione geologica del complesso Diego de Almagro, che favoriva la formazione di centri monogenetici piuttosto che lo sviluppo di vulcani centrali, limitando così lo stazionamento dei magmi basici nei serbatoi crostali e di conseguenza l'interazione con la crosta.

KEY WORDS: *magma mixing, basalts, andesites, Central Andes, mineral chemistry.*

INTRODUCTION

The Central Andes represent an exceptional example of subduction-related, calcalkaline magmatism in a geological setting characterised by thick continental crust, up to 70-75 km (e.g. Beck *et al.*, 1996; Giese *et al.*, 1999; Yuan *et al.*, 2002). In this scenario, most workers agree that the prevailing crustal signature of Central Andes Volcanic Zone (CVZ) magmas is due to mixing with crustal melts during mafic magma ascent and storage (e.g. Davidson and de Silva, 1992; Trumbull *et al.*, 1999; Schmitt *et al.*, 2001; Caffè *et al.*, 2002; Figueroa *et al.*, 2005). The mantle-derived magmas are preferentially stored at the base of the crust or in crustal reservoirs where they differentiate and induce crustal melting, generating high volumes of silicic-intermediate melts (e.g. Hawkesworth *et al.*, 1982; de Silva *et al.*, 2005 and references therein). Between 10 and 4 Ma, several pulses of silicic volcanism consisting of major ignimbrite eruptions characterised the Altiplano-Puna region, a plateau in Central Andes between 21 and 24°S that is considered one of the largest silicic provinces in the world (Lindsay *et al.*, 2001; de Silva *et al.*, 2005). Huge crustal melting is inferred to contribute to silicic volcanism in this region (de Silva, 1989; Coira *et al.*, 1993; Ort *et al.*, 1996), and the heat necessary is attributed to a flux of basaltic magma due to a process of delamination of the lower lithosphere (Kay *et al.*, 1994; Allmendinger *et al.*, 1997; de Silva *et al.*, 2005).

In the Puna region, the Late Miocene volcanism focuses along important NW-SE strike-slip structures running transversal to the arc for hundreds of kilometres toward the hinterland, crossing the Puna-Eastern Cordillera-Subandean-Santa Barbara morphostructural units (Allmendinger *et al.*, 1983; Salfity, 1985; Marrett *et al.*, 1994).

One of these strike-slip structures is the Calama-Olocapato-El Toro (COT) lineament, at 24°S (Fig. 1), along which several important volcanic centres aligned during Miocene from the arc to the Eastern Cordillera (Matteini *et al.*, 2002a, 2002b; Petrinovic *et al.*, 2005).

Along the Calama-Olocapato-El Toro fault system, at the transition between the Puna and the Eastern Cordillera, two volcanic complexes formed during Late Miocene, named Negra Muerta and Diego de Almagro. The easternmost of these, the Diego de Almagro volcanic complex, is constituted by sub-horizontal successions of andesitic lava flows and domes and by minor pyroclastic deposits (Krallmann, 1994; Hauser *et al.*, 2005; Matteini *et al.*, 2005). These rocks, consisting mainly of basaltic andesites to andesites, have a low crustal imprint compared to the typical Mio-Pliocene andesite rocks outcropping in the Puna-Altiplano region. In this paper, we report petrography and mineral chemistry data of selected basaltic andesite rocks belonging to the Diego de Almagro volcanic complex. Detailed petrography and mineral chemistry identified different populations of crystals, revealing that these rocks preserve in their mineralogical assemblage phases in equilibrium with mafic melts and thus may allow to make inferences on the mantle source characteristics in this region.

GEOLOGICAL BACKGROUND

The Central Andes are crossed, between 22 and 27°S, by trans-arc strike-slip fault systems with NW-SE strike, extending inland for several hundreds of kilometres and accompanied by mainly Miocene volcanic activity. The Diego de Almagro volcanic complex (Fig. 1) represents the easternmost Miocene magmatic occurrence along one of these structures, the Calama-Olocapato-El Toro fault system (Krallman, 1994). The complex is located at about 24°S and 66°W in the Eastern Cordillera, at 300 km from the arc and 600 km from the trench (Fig. 1). The adjacent Puna region (Fig. 1), a 4.5 km high plateau, is characterized during Miocene by huge stratovolcanoes and caldera structures. In contrast, the Diego de Almagro complex consists of several, often monogenetic, eruptive centres, aligned along fault systems, producing small volume eruptions

and without the development of a central edifice. In the northern sector of the Diego de Almagro volcanic complex, andesitic to dacitic lava flows and lava domes overlay discordantly a Miocene monzodiorite intrusive body (Hongn *et al.*, 2002; Matteini *et al.*, 2004). The southern sector (Fig. 1) consists of exogenous lava domes, lava flows and minor pyroclastic deposits ranging in composition from basaltic andesites to andesites, overlaying discordantly the Precambrian metasedimentary basement. Recent geochronological datings (Matteini *et al.*, 2005) furnish for the Diego de Almagro complex K/Ar ages between 12 and 7 Ma. Based on petrographic evidence, two groups of andesitic lavas are present (Hauser, 2005). The first group shows amphibole as the main mafic mineral phase, while the latter group, only occurring in the southern Diego de Almagro sector, is characterised by an anhydrous paragenesis with ortho- and clinopyroxene and olivine besides plagioclase. Petrographic and geochemical data on the whole Diego de Almagro volcanic complex can be found in Mazzuoli *et al.* (submitted).

STUDIED MATERIALS

The Diego de Almagro volcanic rocks are subalkaline (Fig. 2) and for their FeO^*/MgO and K_2O values can be classified as HK calcalkaline (Table 1). They range in composition from basaltic andesites to dacites (Matteini *et al.*, 2005). This work is focused on the study of samples of the least differentiated rocks in the Diego de Almagro complex, the basaltic andesites (Table 1), belonging to a scoria cone and to the associated lava flows occurring in the southern sector of the Diego de Almagro complex and dated at about 7 Ma (Matteini *et al.*, 2005). These rocks, having Mg_{value} between 71 and 75 (calculated assuming $\text{Fe}_2\text{O}_3/\text{FeO}=0.15$), represent the most mafic products erupted during Miocene along the entire Calama-Olocapato-El Toro structure, thus they deserve detailed investigations. They are porphyritic rocks, with phenocrysts abundance between about 20 and 30 vol%. At the hand-scale, both the scoria and the lava samples show wave-form bands with different macroscopic colour, and greenish rounded

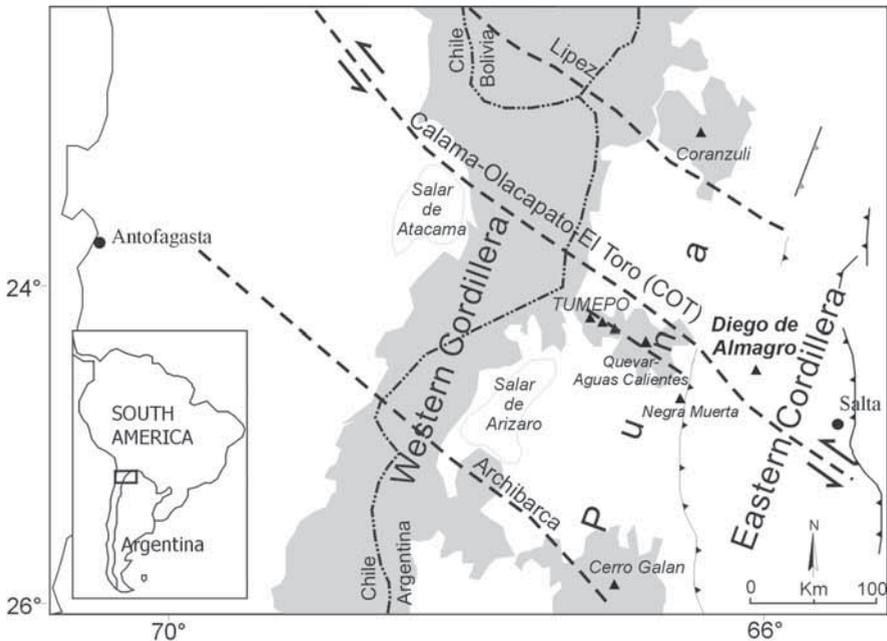


Fig. 1 – Sketch map of Central Andes between 22 and 26°S, showing the location of the Diego de Almagro volcanic complex. Dashed lines: main strike-slip trans-arc structures. Grey area: Miocene-Quaternary arc and trans-arc magmatism.

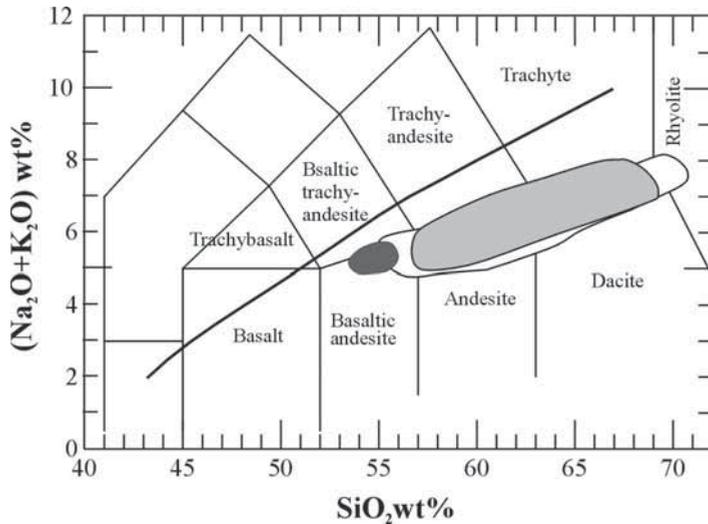


Fig. 2 – TAS diagram (Le Bas *et al.*, 1986) showing the composition of the basaltic andesites of the Diego de Almagro volcanic complex (dark grey field), compared with the composition of the rest of the Diego de Almagro volcanic rocks (light grey field, Matteini *et al.*, 2005) and with the volcanic rocks erupted along the COT during Miocene (open field, Matteini *et al.*, 2002a,b, Petrinovic *et al.*, 2005). The Irvine and Baragar (1981) line dividing the alkaline and subalkaline fields is shown.

TABLE 1 – *Representative bulk rock major elements analyses of basaltic andesite to dacite samples from the Diego de Almagro complex. $Mg_{value} = Mg/(Mg+Fe^{2+})$, with $Fe^{3+}/Fe^{2+}=0.15$. Major oxides were obtained by XRF on fused samples with an ARL 9400 XPb instrumentation at Dipartimento di Scienze della Terra, University of Pisa. Accuracy was monitored using international standards and it is better than 10% for concentrations higher than 10 ppm.*

sample	N-32	N-34	N-52	TA-412	TA-414	TA-113
latitude S	24°24'27"	24°24'27"	24°24'27"	24°22'51"	24°22'51"	24°21'15"
longitude W	65°52'29"	65°52'29"	65°52'29"	65°51'32"	65°51'32"	65°52'06"
SiO ₂ (wt%)	53.17	54.21	57.89	59.54	61.10	66.03
TiO ₂	1.14	1.07	0.98	0.84	0.77	0.42
Al ₂ O ₃	15.02	14.84	17.26	15.95	16.46	16.76
Fe ₂ O ₃	9.24	8.71	7.17	6.39	5.92	3.33
MnO	0.14	0.14	0.23	0.12	0.11	0.10
MgO	5.99	6.78	3.34	3.14	2.49	1.02
CaO	8.01	7.37	6.36	5.28	5.12	3.07
Na ₂ O	2.57	2.67	2.58	3.18	3.17	3.75
K ₂ O	2.55	2.67	3.40	3.45	3.57	4.02
P ₂ O ₅	0.37	0.34	0.31	0.31	0.30	0.19
L.O.I.	1.39	0.09	1.20	0.38	1.09	0.23
Mg _{value}	71.1	74.7	63.8	65.0	61.4	53.7
FeO*/MgO	1.37	1.14	1.91	1.81	2.12	2.91

enclaves, 1-3 cm large. The enclaves, apparently, do not present any geometrical relationships with the banding.

Several thin sections were prepared for the petrographic and microanalytical study of these features.

RESULTS

Petrographic features

The scoria and lava samples of the Diego de Almagro basaltic andesites present macroscopic bands with different colour, having different texture under the optical microscope. The dark bands are rather dense and microlite-rich, while the clear bands are finely vesiculated and show a glassy groundmass. The phenocrysts are plagioclase, clino- and orthopyroxene and olivine. The same mineral phases occur as microlites, with a higher abundance of plagioclase. The dark bands show prevailing mafic phenocrysts and minor sieve-textured plagioclase, while the clear bands show unsieved plagioclase and minor mafic crystals. Hydrous phases, such as hornblende and biotite, that are common in Diego de Almagro andesitic to dacitic rocks, are absent in the phenocryst assemblage of the basaltic andesites. Crystals showing disequilibrium features, such as reaction rims of clinopyroxene mantling orthopyroxene and orthopyroxene mantling olivine, occur both in the clear and in the dark bands. Quartz xenocrysts occur in some samples. They commonly show rounded edges and engulfed rims suggesting re-absorption, and sometimes have a reaction corona made of pyroxene.

The cm-size greenish enclaves visible in the hand-samples are composed of clinopyroxene, mainly showing wavy extinction. Rare phlogopite and orthopyroxene and small grains of iddingsitic olivine occur. The contacts between the enclaves and the host basaltic andesite are linear, regular and well defined.

Microanalytical data

The microanalytical study evidences that phenocrysts of olivine Fo_{75-70} , high- Mg_{value} clinopyroxene with $Cr_2O_3 > 0.5wt\%$ and labradoritic plagioclase are more abundant and appear

prevalently in equilibrium in the dark microlite-rich bands, while in the clear glass-rich bands the prevailing crystals are andesine plagioclase, orthopyroxene, low- Mg_{value} clinopyroxene, Fe-Ti oxides and apatite (Table 2). Phenocrysts commonly show compositional zoning, highlighted by backscattered electrons imaging (Fig. 3a, b; Table 2). Direct zoning of plagioclase (e.g. An_{63} in the core to An_{36} in the rim) characterises the clear bands, while mainly reverse or oscillatory zoning (An_{60} to An_{70}) is common in the dark bands. The wide range in pyroxene composition and zoning is very interesting (Fig. 4). Both clino- and orthopyroxene phenocrysts show reverse zoning (up to Mg_{value} 63 at the core to 84 at the rim in clinopyroxene, Table 2 and Fig. 4), with thick rims ranging mostly from 50 to 200 microns. Somewhere, crystals with opposite zoning trends occur side by side (Fig. 3b, c).

Coarse grains of olivine Fo_{87-83} occur prevalently in the dark bands (Fig. 3d; Table 2d). These olivine crystals include small grains of Al- and Ti-rich diopside averaging $Wo_{47}Fs_{10}$ (Fig. 4) and Cr-spinel ($Cr_{\#}=59-51$, with $Cr_{\#}=Cr/(Cr+Al)$). In one case olivine Fo_{80} contains orthopyroxene Wo_3En_{81} together with Cr-spinel with $Cr_{\#}=59$. These coarse olivine crystals always show a thick iddingsite rim, indicating disequilibrium. The iddingsite alteration is in many cases so developed to impede the analysis of olivine, and prevents to verify the presence of crystal zoning.

The pyroxenitic enclaves are made for the most part of clinopyroxene with composition in the range $Wo_{42}Fs_{11}$ - $Wo_{44}Fs_{13}$ and scarce orthopyroxene $En_{72}Fs_{25}$ - $En_{74}Fs_{21}$ (Table 2c). Interestingly, these pyroxenes have an intermediate Mg_{value} and a lower Cr content if compared to the Mg-rich pyroxene of the dark bands (Fig. 4). In one case we found a granular enclave containing phlogopite with Mg_{value} 74-71 and Al-Fe spinel in addition to ortho- and clinopyroxene.

DISCUSSION

The occurrence in a rock of bands with disequilibrium parageneses and of reaction rim growth on phenocrysts is interpreted as evidence of mingling and mixing of magmas with different composition, and, since the first papers of decades

TABLE 2a, b, c, d, e – Representative analyses of the main mineral phases occurring in the basaltic andesites from the Diego de Almagro complex. bdl: below detection limit. a) Plagioclase. b) Clino- and orthopyroxene in the dark and light bands and in Fo+Cr-spinel+Di clusters. $(Fe/Mg)_{melt}^*$: calculated according to Sisson and Grove (1993). c) Clino- and orthopyroxene in the pyroxenitic enclaves. d) Olivine. e) Fe-Ti oxides recalculated according to Stormer (1983). Major and minor element microanalysis of mineral phases were carried out on polished thin sections with a Philips XL30 scanning electron microscope equipped with microanalysis EDAX (standardless software DXi4) at Dipartimento di Scienze della Terra, University of Pisa (acceleration voltage 20 kV, beam current 5 nA, live time 100 s). The accuracy is lower than 0.5% if abundance is >15wt%, is 1% if abundance is around 5wt% and is lower than 20% if abundance is around 0.5wt%.

TABLE 2a

mineral sample	PLAGIOCLASE							
	N34			N34			N34	
	phenocryst, dark bands			phenocryst, light bands			phenocryst, light bands	
	core	interm.	rim	core	interm.	rim	core	rim
SiO ₂ wt%	52.01	49.57	49.82	47.91	52.2	50.13	51.92	59.36
Al ₂ O ₃	29.03	31.62	31.25	32.8	29.75	30.9	30.16	25.23
FeO	1.38	0.78	0.88	0.76	0.75	0.89	0.77	0.91
CaO	11.93	14.33	14.26	15.67	12.87	14.08	12.83	7.25
Na ₂ O	3.3	3.18	3.12	2.5	4.02	3.58	3.95	6.47
K ₂ O	1.95	0.37	0.4	0.25	0.41	0.32	0.36	0.78
An	58.99	69.82	69.96	76.47	62.38	67.24	62.87	36.46
Ab	29.53	28.04	27.70	22.08	35.26	30.94	35.03	58.87
Or	11.48	2.15	2.34	1.45	2.37	1.82	2.10	4.67

TABLE 2b

mineral sample	CLINO- AND ORTHOPYROXENE								
	N34		N34		N34	N34		N34	
	small grains included in Fo83, dark band		phenocryst, dark band		phenocryst, dark band	phenocrysts, light band			
		core	rim	unzoned	core	rim	core	rim	
SiO ₂ wt%	47.13	47.89	50.37	52.12	52.15	52.65	53.21	52.62	50.65
TiO ₂	2.14	1.81	0.58	0.51	0.42	0.16	0.42	0.33	0.8
Al ₂ O ₃	7.68	7.06	3.36	2.7	2.39	0.88	1.6	1.51	3.32
FeO	6.14	6.26	12.14	5.48	5.51	23.01	17.59	9.94	8.86
MnO	bdl	bdl	0.4	0.24	0.27	0.93	0.61	0.48	0.27
MgO	14.29	14.16	13.17	16.35	16.27	21.75	24.96	14.42	14.73
CaO	21.72	21.92	19.47	21.62	21.89	0.42	1.5	20.25	21.05
Na ₂ O	0.23	0.4	0.37	0.21	0.34	0.19	0.1	0.44	0.32
K ₂ O	0.65	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl
Cr ₂ O ₃	bdl	0.5	0.14	0.77	0.76	bdl	bdl	bdl	bdl
Mg _{value}	80.58	80.13	65.92	84.18	84.04	62.76	71.67	72.12	74.77
Wo	46.81	47.13	41.19	44.44	44.83	0.86	3.00	42.12	43.44
En	42.86	42.37	38.77	46.77	46.37	62.22	69.52	41.74	42.30
Fs	10.33	10.51	20.04	8.79	8.81	36.92	27.48	16.14	14.27
$(Fe/Mg)_{px}$	0.24	0.25	0.52	0.19	0.19	0.59	0.39	0.39	0.34
$(Fe/Mg)_{melt}^*$	1.05	1.08	2.25	0.82	0.83	2.58	1.72	1.68	1.47

TABLE 2c

mineral sample	CLINO- AND ORTHOPYROXENE				
	CS1	N32	CS1	CS1	N32
	pyroxenite enclave unzoned				
SiO ₂ wt%	49.57	51.97	50.63	50.2	52.51
TiO ₂	0.75	0.5	0.47	0.65	0.32
Al ₂ O ₃	4.67	2.85	3.35	4.11	3.7
FeO	8.33	7.51	8.42	7.63	13.49
MnO	0.22	0.23	0.38	0.32	0.36
MgO	15.22	16.03	14.48	15.91	27.09
CaO	20.65	20.46	21.49	20.46	2.15
Na ₂ O	0.27	0.32	0.33	0.33	0.19
K ₂ O	bdl	bdl	bdl	bdl	bdl
Cr ₂ O ₃	0.18	0.13	0.2	0.25	0.18
Mg _{value}	76.51	79.19	75.41	78.80	78.17
Wo	42.73	42.07	44.57	42.14	4.27
En	43.82	45.87	41.79	45.60	74.83
Fs	13.45	12.05	13.63	12.27	20.90
(Fe/Mg) _{px}	0.31	0.26	0.33	0.27	0.28
(Fe/Mg) _{melt} *	1.33	1.14	1.42	1.17	1.21

TABLE 2d

mineral sample	OLIVINE								
	CS1	CS1	CS1	CS1	CS1b	CS1b	CS1b	N34	N34
	unzoned, with rim in equilibrium				with iddingsite rim				
SiO ₂ wt%	38.18	39.11	39.63	38.65	40.46	40.54	40.26	39.65	37.6
FeO	22.45	16.31	13.99	17.9	13.46	12.44	13.28	15.62	24.3
MnO	0.34	0.35	0.29	0.38	0.23	0.3	0.38	0.34	0.77
MgO	38.87	43.77	45.85	42.63	45.55	46.54	45.79	44.27	37.09
CaO	0.16	0.38	0.23	0.32	0.19	0.14	0.13	bdl	0.24
Cr ₂ O ₃	bdl	bdl	bdl	0.05	bdl	0.04	0.14	0.11	bdl
NiO	bdl	0.09	0.03	0.06	0.13	bdl	bdl	bdl	bdl
Fo	75.3	82.4	85.1	80.6	85.6	86.7	85.7	83.2	72.5

TABLE 2e

mineral sample	<i>Fe-Ti OXIDES</i>				<i>Cr, Al-SPINEL</i>	
	N34	N34	N33	N33	N34	N34
	Magnetite	Ilmenite	Magnetite	Ilmenite	small grains in Fo86 in Fo83	
SiO ₂	bdl	bdl	0.43	0.86	bdl	bdl
TiO ₂	4.3	46.23	14.05	46.88	0.61	1.63
Al ₂ O ₃	2.08	0.6	3.9	2.35	15.46	18.26
FeO _{tot}	91.24	47.65	77.72	45.7	34.02	43.49
MnO	0.59	1.07	0.47	1.05	bdl	bdl
MgO	1.58	4.44	2.94	2.95	8.65	8.5
Cr ₂ O ₃	0.21	bdl	0.49	0.21	41.26	28.11
%mol Usp	0.12		0.42			
%mol Ilm		0.83		0.90		
Mg/Fe+Mg					0.31	0.26
Cr/Cr+Al					0.64	0.51

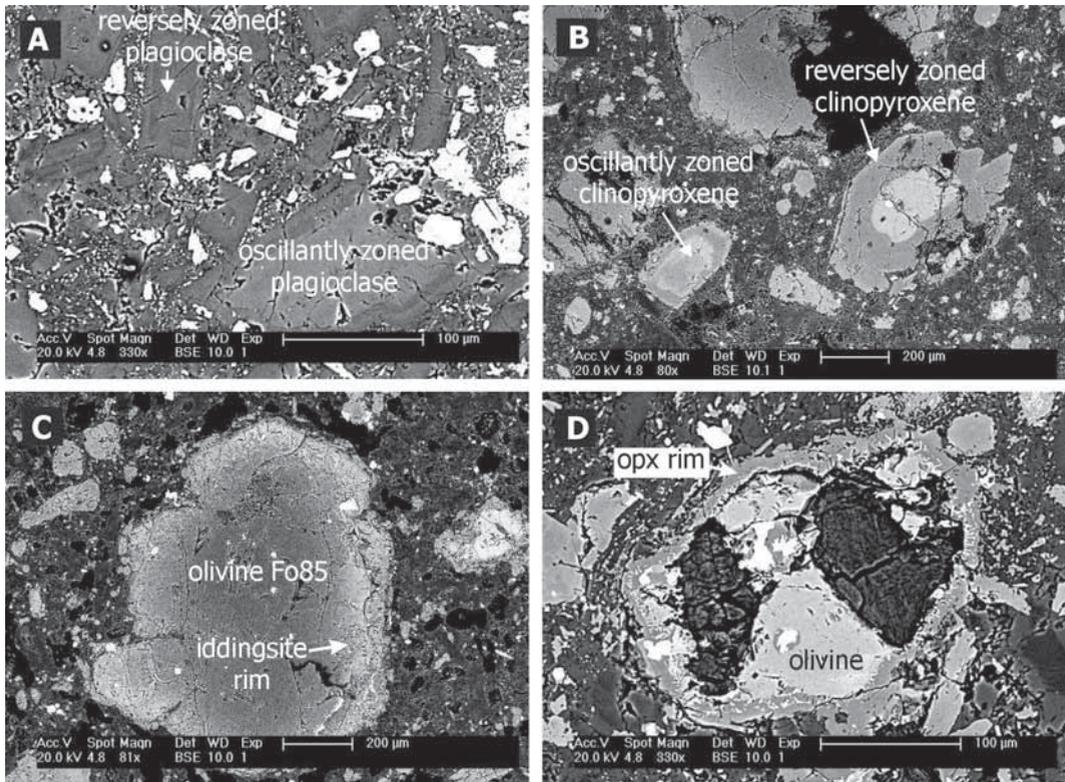


Fig. 3 – Disequilibrium phenocryst assemblages in Diego de Almagro basaltic andesites. SEM backscattered electrons images were collected on polished thin sections with a Philips XL30 scanning electron microscope at Dipartimento di Scienze della Terra, University of Pisa.

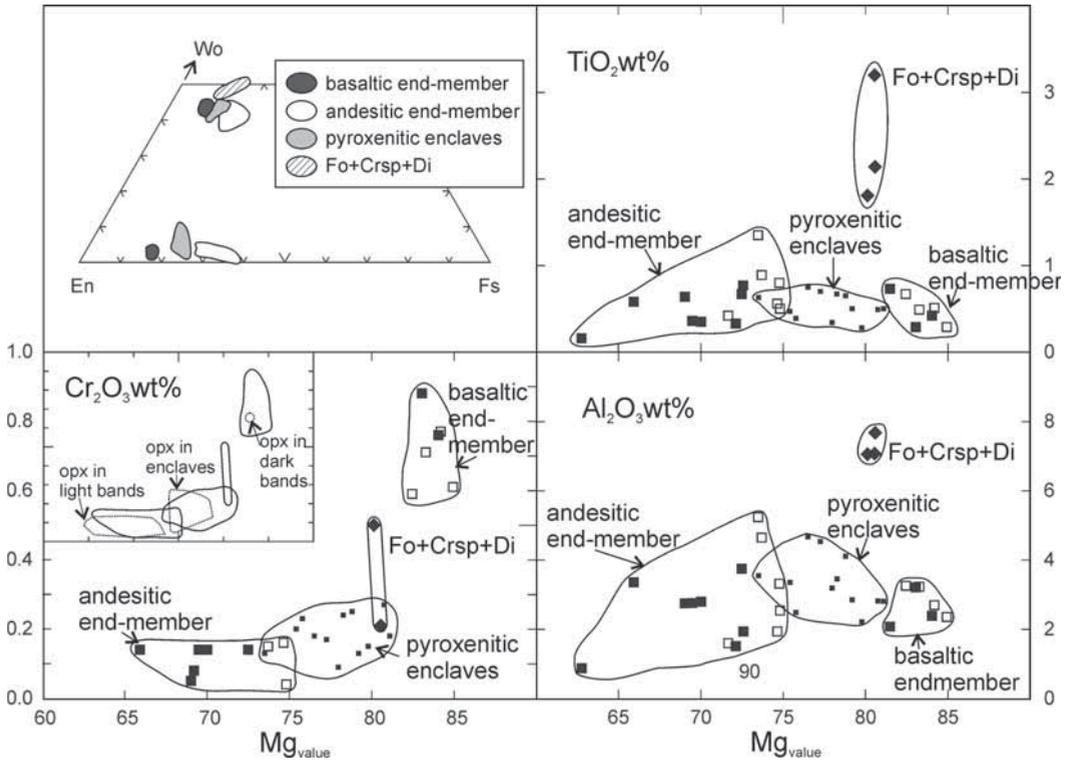


Fig. 4 – a) Wo-En-Fs diagram of clinopyroxene in Diego de Almagro basaltic andesites. b) Cr_2O_3 wt% vs Mg_{value} of clinopyroxene in basaltic andesites (full squares: cores; open squares: rims), their pyroxenite enclaves (small squares) and the Fo+Cr-spinel+Di clusters (diamonds); fields group clinopyroxene corresponding to the basaltic and andesitic end-members of mixing, that do not exactly correspond to the light and dark mingled bands, because crystals of the mafic end-member occur in the light bands and vice-versa. In the inset: the same diagram with dotted fields showing the composition of orthopyroxene in the dark and light bands of basaltic andesites and in pyroxenite enclaves. c, d) TiO_2 wt% and Al_2O_3 wt% vs Mg_{value} of clinopyroxene in basaltic andesites; symbols and fields as in a).

ago (e.g. Eichelberger, 1975, 1978; Bacon, 1986), a large literature developed on the subject both in volcanic and intrusive rocks (e.g. Snyder, 2000; Perugini and Poli, 2000).

The petrography and mineral chemistry of the Diego de Almagro basaltic andesites suggests that the dark bands correspond to a magma with a more mafic composition than the clear bands. The extent of mingling is higher than the simple mechanical formation of bands, because crystals in disequilibrium occur both in the clear and dark bands. In fact, the occurrence and composition of crystal zoning (Fig. 3a, b; Table 2) indicates that crystallisation continued after mingling. Thus, we

may infer that the two magmas underwent to an intimate mixing process, that led crystals in contact with a melt in disequilibrium. The bands visible at the macro-scale are the portions of the rock that better maintain the original composition of the two end-members, but none of them represent the pure end-members.

The mixing process is highlighted by the compositional range of pyroxene (Fig. 4). To better constrain the composition of the two mixing magmas, the Fe^*/Mg ratio ($\text{Fe}^* = \text{Fe}^{2+}$ as total iron) of the liquid in equilibrium with the different populations of clinopyroxene was calculated using a $K_d = (\text{Fe}/\text{Mg})_{\text{px}} / (\text{Fe}/\text{Mg})_{\text{melt}}$ of

0.23, as suggested by Sisson and Grove (1993) for moderate Al clinopyroxenes in calcalkaline rocks (Table 2b, c). The Mg- and Cr-rich cores in the mafic end-member of mixing have Mg_{value} between 82 and 85, and are in equilibrium with a melt with $Fe^*/Mg = 0.8-0.9$, while the cores of crystals in the andesitic end-member have Mg_{value} between 63 and 73 and crystallised in equilibrium with a melt with $Fe^*/Mg = 1.6-2.5$. The rims of pyroxene phenocrysts in the light bands cluster at $Mg_{\text{value}} = 73-75$ (Fig. 4), indicating that they completed their growth in a hybrid melt with $Fe^*/Mg = 1.6-1.9$. The whole rock analyses of the basaltic andesites show FeO_{tot}/MgO ranging between 1.1 and 1.4, thus intermediate between those of the two end-members.

The presence of coexisting ilmenite-magnetite pairs in the clear bands allowed to determine the temperature and fO_2 conditions during crystallisation (Table 3), that resulted close to $\Delta NNO+2$, indicating rather oxidizing conditions, for a temperature around 800°C. No ilmenite-magnetite pair clearly in equilibrium with the mafic assemblage was found.

The coarse crystals of Fe_{87-80} olivine hosting Cr-spinel and diopside do not correspond to the basaltic mineral association observed prevalently in the dark bands (Fig. 4), thus they can be considered as xenocrysts, possibly derived from early crystallisation of mantle-derived magmas. The forsterite content of olivine, if compared to the $Cr_{\#}$ of coexisting spinel, plots out of the Arai (1994) spinel-forsterite mantle array, and is low to be in equilibrium with mantle peridotite. Nevertheless, the chemistry of olivine, spinel and diopside and the lack of the accessories (apatite, Fe-Ti oxides)

that are common in the Mg-rich clinopyroxene phenocrysts of the mafic end-member indicate that these crystals belong to a mineral association more primitive than that corresponding to the mafic end-member of mixing. The application of the cpx-opx geothermometer to clinopyroxene and orthopyroxene both hosted in Fe_{80} olivine suggests high temperature of equilibrium, about 1350°C, for the most Mg-rich compositions. The Al^{IV} - and Ti-rich and Si-poor character of diopside included in the Fo-rich olivine is peculiar and is not characteristic of the most Mg- and Cr-rich clinopyroxene in the dark bands (Fig. 4). A trend to high Al^{IV} and Ti recalls the compositions typical of alkaline basalts. On the other hand, several experimental studies evidenced that the abundance of these minor components in clinopyroxene may reflect a higher pressure of crystallisation (e.g. Adam and Green, 1994).

The pyroxenite enclaves found scattered in the basaltic andesites show a mineral chemistry which is Mg- and Cr-poor in respect to the basaltic association, but that cannot correspond for texture and mineral composition to cumulates of an andesitic magma chamber (Fig. 4). We interpret these enclaves as deep mafic cumulates, extracted by the rising magma. The application of the cpx-opx geothermometer to the pyroxenitic enclaves indicates a temperature between 1190-1240°C.

The presence of the described disequilibrium paragenesis and the involvement of high-T and possibly high-P mineral phases suggests that the eruption of the basaltic andesites had a complex history, tapping magmas originated at depth together with magmas stored and differentiated at some levels in the crust. Some considerations can be made on the mechanisms of mixing with the andesite magmas.

The different compositional shells in zoned crystals of Diego de Almagro basaltic andesites show a thickness ranging mostly from 50 to 200 microns. Considering a growth rate of the order of 10^{-10} cm/s (Marsh, 1988), several years are necessary for the rims to grow. In undercooling conditions, such as the case of a little fraction of mafic hot magma entering a cooler silicic one, the growth rate may be much higher and similar features may form in nearly syn-eruptive conditions (several days or weeks). However, in

TABLE 3 – Temperature and oxygen fugacity calculated from the composition of five different pairs of coexisting magnetite and ilmenite with the Spencer and Lindsley (1981) Fe-Ti oxide geothermometer and the software ILMAT.

Sample	Temperature (°C)	Log ₁₀ fO ₂
N34	779	-12.51
N34	822	-11.32
N34	754	-12.98
N34	788	-11.66
N33	860	-12.75

our case, although the temperature difference was probably over 100°C, the mass proportions of the basaltic and andesitic end-members should not be very different. Thus, it is no reason to imagine undercooling conditions in our mixing case. Therefore, the time needed for the thickness of the observed overgrowth rims indicates that mixing did not occur in syneruptive conditions, but in the magma chamber, during magma storage in the crust. The extent of blending of basalt and andesite requires that the viscosity contrast was low. Given the compositional and temperature contrast of basalt and andesite, we may infer that the volatiles dissolved in magma played a fundamental role in lowering the viscosity contrast and allowing the intimate mingling observed in the samples.

The main mechanisms driving the formation of disequilibrium paragenesis may be caused either by convective motions of crystals in a thermal and compositional gradient in a magma chamber (Singer *et al.*, 1995; Couch *et al.*, 2001; Ginibre *et al.*, 2002), or by a mafic recharge in an evolving chamber. The presence of the primitive xenocrysts of olivine in our samples may suggest an episode of mafic recharge of a chamber evolving from basalt to andesite. However, the mass proportions of basalt and andesite in the erupted products induce to hypothesise that the eruption emptied a basalt-andesite reservoir, in which mixing occurred before the eruption because the physical properties of the magmas allowed convective motions of crystals and melt.

CONCLUSIVE REMARKS

In a geological setting such as that of the Central Andes, the characteristics of the mantle sources are commonly obscured by the strong crustal contamination affecting the magmas ascending through the thick continental crust. However, the volcanic rocks may preserve traces of their mantle-derived parent magmas in the form of xenocrysts or in diverse populations of phenocrysts coexisting due to magma mixing. In this work, we attempted to find such traces studying some of the least differentiated rocks erupted along the COT system during Late Miocene, occurring in the Diego de

Almagro complex, at the easternmost end of the COT.

The conclusive remarks suggested by the data discussion can be summarised as follows:

– The results of this study revealed that the least differentiated of the Diego de Almagro rocks preserve evidence of the involvement of primitive basaltic magmas in the eruption. The preservation of primitive crystal assemblages was probably favoured by the lack of development of a central edifice, with a stable magma reservoirs where primitive assemblages could be obliterated by the strong interaction with the crust. In the Diego de Almagro region, instead, magmatism occurred by monogenetic volcanic centres.

– The petrographic and chemical data indicate that mixing with andesitic magma occurred during the ascent and storage of the primitive magmas in the crust before eruption. A slightly higher fraction of basaltic magma compared to the andesitic one was involved in the mixing process. Crystal zoning and disequilibrium features suggest that intimate hybridisation occurred, driving crystals to prosecute their growth in melts with different composition.

– The mixing modalities and the presence of high-T mineral assemblages and of pyroxenitic enclaves lead to exclude that the basaltic magma represented a refilling episode of an andesitic chamber stored in the crust. It seems more probable that the eruption emptied a basaltic reservoir with a differentiated andesite cap, in which mixing occurred before the eruption because the physical properties of the magmas allowed convective motions of crystals and melt.

– The data obtained evidence the necessity and effectiveness of detailed petrography and mineral chemistry in the identification of different magmatic components in the Andean rocks. The results of this study shed some light on the complex history of mixing and mingling from the source up to the storage in magma chambers within the continental crust for the Diego de Almagro volcanic rocks. Further investigations of these rocks with advanced analytical techniques, allowing the determination of in-situ trace and isotope characteristics, are needed to unravel the characteristics of the mantle-derived magmas and of the crust with which they interacted.

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